

The Atmospheric Response to Realistic Reduced Summer Arctic Sea Ice Anomalies

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ABSTRACT

The impact of reduced Arctic summer sea ice on the atmosphere is investigated by forcing an atmospheric general circulation model (AGCM) with observed sea ice conditions during 1995, a low-ice year. The 51 experiments, which spanned April to October of 1995 were initiated with different states from a control simulation. The 55-year control was integrated using a repeating climatological seasonal cycle of sea ice. The response was obtained from the mean difference between the experiment and control simulations.

The strongest response was found during the month of August where the Arctic displays a weak local thermal response, with warmer surface air temperatures (SAT) and lower sea level pressure (SLP). However, there is a significant remote response over the North Pacific characterized by an equivalent barotropic (anomalies are collocated with height and increase in magnitude) structure, with anomalous high SLP collocated with a ridge in the upper troposphere. The ice anomalies cause an increase (decrease) in precipitation along (north of) the North Pacific storm track. A Linear Baroclinic Model forced with the transient eddy vorticity fluxes, transient eddy heat fluxes and diabatic heating separately demonstrated that transient eddy vorticity fluxes are key to maintaining the anomalous high over the North Pacific.

The model's sensitivity to separate ice anomalies in the Kara, Laptev-East Siberian, or Beaufort Seas includes SLP, geopotential height, and precipitation changes that are similar too but weaker than the response to the full sea ice anomaly.

1.0 Introduction

Summer sea ice in the Arctic decreased at a rate of 4-6% per decade (Deser et al. 2000) through the 1990's and the melt rate has accelerated in the 2000s when the rate of decline has increased to 10% per decade (NSIDC 2007, Stroeve et al. 2005). In the 1990's the melting of Arctic ice was consistent with the positive phase of the North Atlantic Oscillation (NAO), which is characterized by enhanced storminess and warm moist air penetration into the Arctic. The enhanced moisture in the Arctic has increased downward longwave fluxes to hasten ice retreat in summer (Francis and Hunter 2006). The NAO has approached more neutral values since 2000 yet the ice melt has accelerated. The observed influx of plugs of warm Atlantic layer water into the Arctic provides one likely mechanism for the recent ice melt (Polyakov et al. 2005), though the exact process by which heat from the Atlantic layer reaches the upper ocean is not presently understood. The summer sea ice is expected to continue its decline based on the recently documented decreases in winter ice (Comiso 2006) and the warm Atlantic Water headed for the Arctic that is being tracked by various ocean observing programs (Polyakov et al. 2007). In a warmer climate large decreases in summer sea ice may be more common and while the ice anomalies initially result from both atmospheric and oceanic forcing, we hypothesize that they can in turn markedly alter the air-sea exchanges of heat and moisture to subsequently influence the large-scale climate.

The summer warming and sea ice reductions are correlated with cold season circulation changes (Wallace et al. 1996), which lead to changes in advection. For example, reduced summer sea ice in Barents-Kara-Laptev is associated with anomalously low pressure centered in the Arctic the preceding spring (April-June)

(Deser et al. 2000). The general tendency towards lower pressure in the Arctic (Walsh et al. 1996) from 1960-2000 is consistent with enhanced penetration of storms into the Arctic. There has been an increase of warm season cyclone count and intensity in the Arctic (north of 60N) since late 1950's (Serreze et al. 1997, Zhang et al. 2004). Maslanik et al. (1996) find an increase of cyclone activity over the central Arctic Ocean, which advects warm southerly winds into the Laptev and E. Siberian Seas as well as transports ice away from the coast. Ice that is particularly thin as a result to wintertime circulation patterns can be easily broken down and transported due to summer storms, further reducing ice area/concentration.

When high albedo ice is replaced by low albedo ocean, there is significantly more net solar flux at the surface, increasing the heat stored in the upper layer of the ocean. This heat stored during the summer can then slow the freeze up the following winter as well as melt ice at the ice/ocean interface. There is still a reasonably strong correlation (+0.6) between sea ice during the summer and that the following winter (Deser et al. 2000). In the marginal seas when the ice retreats the SST warms by several degrees by August. Fluxes of sensible and latent heat into the atmosphere increase with a warmer ocean, which we hypothesize, could exert some influence back on the atmosphere.

Climate models are ideal tools for understanding the influence of sea ice on the atmosphere because in the observations, climate anomalies are dominated by the atmospheric forcing of the ice. Singarayer et al. (2006) ran the Hadley Centre Atmospheric Model (HadAM3) with climatological SSTs and observed sea ice concentrations from 1978-2000. The simulations were continued from 2001-2100 using sea ice from a moderate (EXP1) and a more severe future projection (EXP2) to

construct a suite of simulations from 1978-2100. Model SAT anomaly tendencies match observations best over the 1993-1995 period (See Figure 4a in Singarayer et al. 2006) when sea ice anomalies display large interannual variability and reached a low for the decade in 1995. They argue that the ice anomalies were likely large enough that sea ice forcing dominated the atmospheric response. The winter response is characterized by warmer SATs and increasing precipitation rates over the Arctic and a weaker (stronger) storm track over the Pacific (Atlantic). The SAT response during summer to only the sea ice anomalies was weak but becomes statistically significant when observed above normal SSTs replace climatological values. Sewall (2005) investigated the response to reduced Arctic sea ice in Community Climate Model Version 3 (CCM3) and a suite of coupled IPCC Fourth Assessment simulations and found a robust pattern of reduced wintertime precipitation for the Western US by ~30%.

Magnusdottir et al. (2004) and Deser et al. (2004) investigated the response to sea ice and SST anomalies during winter in the North Atlantic using CCM3. The ice anomaly pattern corresponds to an enhanced observed trend with ice reductions (increases) east (west) of Greenland. Magnusdottir et al. (2004) found a significant model circulation response to sea ice that resembled the negative phase of the North Atlantic Oscillation, which is opposite of the atmospheric pattern that forced the observed sea ice trend, suggesting that sea ice has a negative feedback on the atmosphere. There is growing evidence that a model's internal variability influences its forced response. To investigate this further, Deser et al. (2004) decomposed the atmospheric response to sea ice into the part that projects on the leading mode of model variability and the residual from this projection. The leading mode has an equivalent barotropic vertical structure and

resembles the NAO, while the residual is baroclinic. A subsequent study by Deser et al. (2007) examines the transient response to wintertime sea ice anomalies in the North Atlantic. They analyzed the GCM output using a Linear Baroclinic Model (LBM) to show that the initial local response is baroclinic and forced by the diabatic heating anomalies associated with surface heat fluxes resulting from reduced sea ice area. The equilibrium response is large scale in extent and barotropic and is largely maintained by the transient eddy vorticity fluxes. Peng and Whittaker (1999) elucidated this eddy driven mechanism to describe the atmospheric response to midlatitude SSTs in an idealized GCM, which can be applied to surface changes resulting from decreased sea ice. These studies show that the atmosphere responds to surface boundary conditions in ways that can influence the storm track.

Alexander et al. (2004) forced CCM3 with realistic sea ice conditions from 1982-83 that had a similar pattern but with a smaller ice area than the anomalies from Magnusdottir et al. (2004) and Deser et al. (2004). The pattern of response is similar in the three studies, with positive (negative) height anomalies in the Arctic (midlatitudes). A comparison of ice area to the strength of 500-hPa response reveals a linearly increasing relationship (See Figure 9, Alexander et al. 2004).

Alexander et al. (2004) also examined the response to ice anomalies in the North Pacific and found that the atmospheric response suggested a positive feedback of the ice on the atmosphere. The different atmospheric responses to ice in the North Atlantic and North Pacific may arise from the position of the storm track relative to the ice edge. In the North Atlantic the ice edge is in the vicinity of the storm track whereas in the North Pacific the ice edge is well north of the storm track. A thorough discussion of

additional studies of the response to winter sea ice is presented in Alexander et al. (2004).

Numerous GCM simulations have investigated the impact of winter sea ice on the atmosphere but few have examined the atmospheric response to sea surface temperature (SST) or sea ice during the summer months. Several studies find the response during summer to be much weaker than winter and focus their analysis on winter (Parkinson et al. 2001, Singarayer et al. 2006). Raymo et al. (1990) reduced the ice to paleoclimatic conditions throughout the year that reached an ice-free Arctic during the month of September. During JJA they found a 3°K warming over Greenland and an overall warming over the polar region. They found no significant differences in sea level pressure, evaporation/precipitation ratios, or cloudiness in the North Atlantic.

This study focuses on the atmospheric response to reduced realistic summer sea ice in the Arctic from the summer of 1995, which had the lowest June-September ice area (based on both extent and concentration) with the exception of the summer of 2007 (Figure 1). In addition to using realistic sea ice extents and concentrations in the Arctic, the other unique features of our study include the summer focus and the use of a large (51) number of ensemble members for each set of experiments to enhance the signal to noise ratio. In addition, a suite of further experiments is conducted to investigate the sensitivity of the model to the location of the ice anomaly and a Linear Baroclinic Model (LBM) is used to diagnose the forcing to assist in the interpretation of the results.

Some key questions that we will address in this study:

- Does the northern hemisphere atmosphere respond to realistic summer time Arctic sea ice anomalies? Is there a remote response as well as a local response? How does the response during summer differ from winter?
- Is the atmospheric response sensitive to the placement (latitude/longitude) of the summer Arctic sea ice anomalies?
- How does the response to sea ice extent compare with that to concentration?

The model experiments are described in Section 2 and the results in Section 3. The summary and a discussion of mechanisms is presented in Section 4.

2.0 Model Experiments

a. Boundary Conditions and Experiment Design

Boundary conditions for the simulations are from the Hadley Center ice concentration and sea surface temperature data set (HadISST version 1.1 - Rayner et al. 2003) and climatologies are based on the 1979-99 period. Observed monthly mean values were interpolated to the model grid using bilinear interpolation over the open ocean and by averaging nearby grid values in coastal regions. Arctic sea ice area varies while thickness is specified to be 2.5 m thick. It is not expected that specifying ice thickness will significantly influence the results since the summer atmosphere in a regional climate model was shown to be insensitive to changes in sea ice thickness (Rinke et al. 2006). Global SSTs and sea ice in the Southern Hemisphere (specified to be 1 m) evolve according to the mean seasonal cycle in order to isolate the influence of Arctic sea ice. In regions where the ice extent was lower than the mean ice cover, the exposed ocean was set to the climatological SST; when the ice area expanded above normal, SSTs

were blended from -1.8°C (the temperature at which there is 100% ice cover) at the ice edge with climatological values from two grid boxes (2.8° latitude x 2.8° longitude) seaward from the ice edge. This method was employed to smooth the temperature gradient between ice and ocean. In the extent experiments, the monthly Arctic sea ice values were specified to cover 100% of the grid square if the observed monthly averaged concentration exceeded 15%; otherwise the grid square was set to be ice-free. Monthly mean ice and SST values were linearly interpolated in time from mid-monthly values to obtain smoothly varying daily extents and concentrations. As a result, the transition from no ice to complete ice cover in a grid square is not instantaneous in the extent simulations, instead the amount of ice linearly evolves between 0% to 100% within the 30-day period when ice forms or melts. While this provides for a smooth transition of the ice edge in space and time, and is probably more realistic than an instantaneous transition, it also introduces fractional ice cover into the extent experiments.

b. Experiment Design

We focus this study on the summer of 1995, which contained minimum sea ice cover over the entire Arctic from June to August during recent years, with the exception of 2007 (Figure 1). Ice area during 2005 and 2006 was just slightly above that of 1995. Typically Arctic summer sea ice extent reaches a minimum in mid-September of approximately $5 \times 10^6 \text{ km}^2$, though recent minima (e.g. 2005 through 2007) have been consistently lower. Three model experiments have been performed in which Arctic sea ice varies according to observations:

- Ice extent varies over April to October of 1995 (Sum95e)

- Ice concentration varies over April to October of 1995 (Sum95c),

where the experiments are designated, in parentheses above, by the season, year and ice configuration. We also performed an extended (55 year) control simulation in which ice extent repeats the same seasonal cycle each year based on the average of the 1979-99 period (Cntle),

The Sum95e and Sum95c experiments consist of an ensemble of 51 CCM3 simulations that extend from April to October. Each member of the ensemble is initialized from a different April 1 from year 5-55 of the control extent experiment (Cntle). A discussion quantifying the relationship between signal to noise ratio and ensemble members is given in Alexander et al. (2004). The modeling results are generally presented as monthly anomalies constructed by averaging over the 51 ensembles and subtracting the corresponding long-term monthly mean over the last 51 years of the control simulation.

The discussion in this paper focuses on the model response during August of 1995. Ice anomalies evolve during the simulation based on observed April to October ice conditions. In 1995, ice was below normal in the Kara-Barents Sea in July and throughout the Eurasian Arctic, the Chukchi and Beaufort Seas during August (Figure 2) and September. The atmospheric response in June and July was generally weak and will not be discussed. This may be a consequence of overall smaller sea ice anomalies during these months.

c. Atmospheric General Circulation Model

The CCM (version 3.6) is the atmospheric GCM used in this study and has 18 vertical levels and a horizontal spectral resolution of T42, which is approximately 2.8° latitude by

2.8° longitude. Kiehl et al. (1998) described the model physics, while Hack et al. (1998) and Hurrell et al. (1998) evaluated the model's climate. While the model has some deficiencies over the Arctic, e.g. it's colder and wetter than observed (which also occurs in most other AGCMs [Randall et al. 1998]), many aspects of the earth's climate are well simulated. This is a well-documented model that has been used in numerous studies of the impact of sea ice on the atmosphere (e.g. Deser et al. 2004, Magnusdottir et al. 2004, Alexander et al. 2004).

d. Linear Baroclinic Model

To understand the mechanism for the large-scale response over the North Pacific to reduced Arctic sea ice in August, we forced a linear baroclinic model (LBM) with daily mean diabatic heating and transient eddy heat and vorticity fluxes, similar to Deser et al. (2007), from Cntle and Sum95e. The LBM (see Peng et al. 2003) is based on the primitive equations configured with T21 horizontal resolution and 10 equally spaced pressure levels from 950-hPa to 50-hPa. The model is linearized about the CCM3 basic state obtained from the long-term August mean in Cntle. The LBM includes dissipation in the form of Rayleigh friction in the momentum equation and Newtonian cooling in the thermodynamic equation, as well as biharmonic thermal diffusion. The Rayleigh and Newtonian damping timescales are 1 day at 950-hPa transitioning linearly to 7 days above 700-hPa. The LBM is integrated for 31 days.

The pattern of the CCM3 response to sea ice forcing is diagnosed by comparing the LBM responses to anomalous diabatic heating and transient eddy fluxes from Cntle and Sum95e. The transient eddies are based on a 14-day high pass filtered data,

constructed by subtracting the 11-day running means from the raw daily data (the half-power point of this filter is 14 days).

3.0 Results and Discussion

3.1 Local Arctic Response

The model atmosphere displays a local thermal response to reduced western Arctic sea ice extent. The net heat flux anomalies resulting from the reduced sea ice are 10-25 W m^{-2} from the ocean to the atmosphere (Figure 3a). The latent heat flux is the dominant form of heating contributing about 4-8 W m^{-2} , followed by sensible heat flux at 2-6 W m^{-2} and then longwave at 2-4 W m^{-2} . Increased upward (downward) directed longwave radiation of 2 W m^{-2} is associated with a decrease (enhanced) in low-level clouds of 2%. Plots of model net surface solar flux, albedo, surface temperature, cloud cover and specific humidity are described in the text but will not be shown. Anomalies of net surface solar heat flux are directed into the ocean and are on the order of 15 W m^{-2} where high albedo ice is replaced by a lower albedo ocean. However, the shortwave anomalies do not impact our simulation since the ocean temperature and ice are fixed and are not included in the net heat flux calculation. In nature, the enhanced solar flux into the surface would melt more ice or act to warm the ocean in the shallow ice-free seas. However, we specify the observed evolution of sea ice and argue that any ice melt from increased net solar radiation into the ocean is represented by the observed sea ice conditions.

The surface temperature anomalies associated with the reduced ice area are between 0.5-1.5°C where climatological ocean sea surface temperatures replace sea ice. The surface air temperature (Figure 3b) warms throughout the Arctic with strongest

warming present over the Kara-Barents and E. Siberian Seas and over eastern Siberia with anomalies between 0.5-1.5°C. This low-level warming is associated with small decreases in sea level pressure and geopotential heights that are not statistically significant (Figure 4). The air temperature warming is relatively shallow over the Arctic with no significant anomalies at or above 925-hPa. There is a significant increase in convective precipitation (1 mm day^{-1}), convective clouds (1-2%) and middle-level clouds (2-4%) in the Laptev Sea where sea ice is reduced. In the Kara Sea over the reduced ice cover, there is a significant decrease in total cloud cover (2-4%), which results from less low, medium and high clouds. There are no significant changes in large-scale precipitation over the Arctic.

3.2a Midlatitude Response

The model response to reduced western Arctic sea ice in the North Pacific is characterized by changes in the large-scale circulation and displays significant anomalies in Eastern Siberia (65°N 165°E) and over the ocean storm track region (55-60°N). In far Eastern Siberia, negative sensible and longwave heat flux anomalies total 5 W m^{-2} (Figure 3a) while downward solar heat flux is reduced by 5 W m^{-2} , resulting in a net heat flux of near zero. Surface temperature and surface air temperatures are warmer by up to 1.0 and 0.5°C (Figure 3b), respectively. Increased convective and large scale precipitation is collocated with increases in specific humidity (up to 2 g kg^{-1}). Total cloudiness in Eastern Siberia increases by up to 4% with more clouds at all levels. The SLP and geopotential height anomalies are weakly negative and not significant over Eastern Siberia. Since the net surface heat flux anomalies are weak, the warmer

moister atmosphere results from southerly advection associated with the circulation changes over the North Pacific.

The SLP response is characterized by a significant anomalous high over the North Pacific with a central maximum of 2-hPa (Figure 4a). At 500 and 200-hPa the anomalous high in the North Pacific reaches 20 and 30 m (Figure 4b and 4c), respectively, displaying an equivalent barotropic structure. This pattern is characteristic of the equilibrium response to a midlatitude heating source attributed to transient eddy feedbacks that results from the interaction of the forced anomalous flow and the storm tracks (Kushnir and Lau 1992; Ting and Peng 1995; Peng and Whittaker 1999). EOF 1 of August SLP in the North Pacific sector in Cntle explains 34% of the variance and displays an anomaly of one sign in the Arctic and of opposite sign in the North Pacific in the area of the model response but with a slightly more zonally elongated shape. The similarity between the EOF pattern and the forced response to reduced sea ice is consistent with the notion that a model's forced response projects onto its intrinsic modes of variability.

The model displays a response in the North Pacific storm track region with significant total precipitation anomalies (Figure 5). Anomalies of large-scale precipitation are about twice as large as those of convective precipitation. The magnitude of the total precipitation response reaches values of 25% of the mean climatological precipitation (Figure 5, contours). The mean and anomalous precipitation patterns suggest a weakening of the main North Pacific storm track and a slight enhancement on the poleward side. In other words, the storm track shifted northward and weakened.

Storm track variability as indicated by 2-8 day bandpass filtered variance statistics for 500-hPa height variance, 850-hPa transient heat flux, 200-hPa transient momentum flux and 500-hPa omega variance are shown in Figure 6. All of the mean model storm track measures in the control simulation (Cntle) display maxima in the eastern Pacific, eastern North Atlantic and in central Eurasia (Figure 6, contours). Decreased ice (Sum95e) leads to a general weakening of the storm tracks throughout the hemisphere (Figure 6, blue shading) where the primary significant response is in the Pacific Sector. There is a small region of enhanced storm track activity over the far Eastern Siberia-Bering Sea region (Figure 6, red shading). Note that an increase in 500-hPa height variance signifies both the passage more highs as well as lows. The bandpassed 500-hPa omega vertical velocity variance anomalies (Figure 6d) are consistent with the height variances. The 2-8 day bandpassed 850-hPa $v'T'$ or transient eddy heat fluxes are reduced over the mean storm track in the North Pacific and enhanced to the north in eastern Siberia-Bering Sea. In addition, the transient eddy heat fluxes at 850-hPa display significant reductions in storm track activity over North America into the North Atlantic. The 2-8 day bandpassed 200-hPa $u'v'$ transient eddy momentum fluxes are characterized by increased (decreased) poleward momentum flux to the north (south) of the mean storm track in the North Pacific, which is consistent with the anomalous high in geopotential height response (Figure 6c). Referring to the geopotential tendency equation, the convergence of vorticity (or momentum) fluxes north of the mean storm track is consistent with the positive equivalent barotropic height anomalies (Lau and Nath 1991). The significant precipitation anomalies (Figure 4) are consistent with the weakening and northward displacement of the North Pacific storm track (Figure 5).

3.2b Diagnosis of Forcing

One possible mechanism for the remote response over the North Pacific involves a Rossby wave train (albeit weak in this case) that is initially excited by diabatic heating anomalies in the Arctic. This wave train propagates into the North Pacific, where through interactions with the storm tracks, changes the storm track to generate an anomalous high over the center of the basin. This mechanism resembles the large scale-eddy feedback described in Peng et al. (2003) with the exception that the boundary forcing was close to the storm track in their study.

Diabatic heating anomalies are constructed to investigate the forcing of the atmosphere by reduced sea ice cover. The Cntle mean vertically integrated diabatic heating is shown by contours in Figure 7a and displays cooling of $50\text{-}100\text{ W m}^{-2}$ over the Arctic. The vertically integrated diabatic heating displays positive anomalies where Arctic sea ice is reduced of $15\text{-}25\text{ W m}^{-2}$, which is about 10-20% of mean. There is a decrease in the region of the North Pacific storm track (Figure 7a) A vertical cross section through the largest diabatic heating anomalies indicates that in the Arctic the positive heating anomalies are located below 800-hPa and the negative anomalies in the North Pacific penetrate up to 400-hPa (Figure 7b).

The Linear Baroclinic Model described in Section 2 was forced with the transient eddy vorticity fluxes, transient eddy heat fluxes and mean diabatic heating separately to diagnose the key forcing behind the atmospheric response patterns. The LBM response (Figure 8c-h) is compared to the full GCM anomalies (Figure 8a and b) at 950-hPa and 500-hPa. This diagnostic model analysis reveals that the transient eddy vorticity fluxes are responsible for maintaining the anomalous high in the North Pacific, whereas

transient eddy heat fluxes and diabatic heating yield a negligible response. The LBM response to the total transient eddy and diabatic heating is nearly indistinguishable from the response to the transient eddy vorticity fluxes. The LBM analysis does not reveal how the reduced Arctic ice anomalies induced the eddy momentum fluxes over the North Pacific, perhaps indicating that the diabatic heating anomalies over the Arctic are too shallow and weak to drive the large-scale response.

3.3 Partial Ice Reduction Sensitivity Experiments

To investigate the sensitivity of the atmospheric response to the placement of the ice anomalies, three experiments were conducted where CCM3 was forced with partial sea ice anomalies from Sum95e ice conditions. The ice was removed in the Kara (Sum95ke) (Figure 9a), Laptev-East Siberian (Sum95le) (Figure 9b) and Beaufort (Sum95le) (Figure 9c) Seas. The integration and processing procedure for the partial ice experiments was similar to one used for full anomaly case (Sum95e) to construct a 51 ensemble member response. The largest positive net surface heat flux anomalies (not shown) into the atmosphere are located directly over grid boxes where ice was removed and look indistinguishable from the analogous anomalies from the Sum95e (Figure 3, top panel) simulation. Sum95ke and Sum95le display weak negative heat flux anomalies over Eastern Siberia and Sum95be has significant negative anomalies around 5 W m^{-2} .

Surface temperature and SAT responses to the partial ice anomalies are characterized by warming in the vicinity of the reduced ice anomaly and the magnitudes are nearly identical to those from the full ice experiment. Warm SAT anomalies in far

Eastern Siberia-Bering region are significant in Sum95le and Sum95be, with the Beaufort ice forcing the largest response in Eastern Siberia.

The atmospheric SLP and geopotential height responses to the partial ice experiments resembles the response in Sum95e. A weak high over the Kara Sea, a weak low over East Siberia stretching into the Chukchi Sea, and the anomalous high in the North Pacific are all common features of the SLP and geopotential height response patterns to partial ice anomalies (Fig 9d-i). The individual responses shown in Figure 9 are weaker than the response in Sum95e, however, the sum of the patterns in Figure 9 for SLP and 500-hPa geopotential height is nearly twice as strong as the response to the total ice anomaly (Sum95e). Ice reductions in the Laptev-East Siberian and Beaufort Seas produce a statistically significant response in the North Pacific. The anomalous low (SLP and 500-hPa height) over East Siberia is stronger in the Beaufort partial ice experiment than in Sum95e. This suggests that the model atmosphere is sensitive to ice reductions in all three of these regions and the induced climate anomalies are fairly similar.

The precipitation response (not shown) patterns to the partial ice anomalies are sensitive to the location of the ice anomalies. The positive precipitation anomalies over Eastern Siberia are weakly evident in Sum95le and are significant in Sum95be. The negative precipitation anomalies in the mean model storm track zone are overall largest for Sum95be, largest over south coastal Alaska for Sum95ke and significant for a limited area over the ocean for Sum95le. The sum of the precipitation anomalies for the three partial ice experiments is slightly larger than the precipitation anomalies for Sum95e.

3.4 Ice Concentration Experiments

The August 1995 experiment was repeated using ice concentration anomalies (Sum95c) (Figure 10a). The net surface heat fluxes (not shown) and SAT were similar to Sum95e. One feature different from the Sum95e ensemble average is an area of significant negative surface air temperature anomaly between 120-150E in Eastern Siberia (Figure 10b). The Sum95c SLP response has a weaker anomalous high in the North Pacific and a stronger anomalous low in Eastern Siberia compared to the extent experiment. The atmospheric response at 500-hPa is similar though it looks more like a wave train in the Pacific (Figure 10c). During summer the contrast between using ice extent and concentration is small whereas the differences are larger in winter (Alexander et al. 2004). The area of open water is slightly larger during summer than winter (compare Figure 1 with Figure 1 of Alexander et al. 2004) but the larger air-sea temperature contrast in winter has a strong influence on the turbulent heat fluxes. Anomalies for Sum95c are constructed by taking the difference between the concentration experiment (Sum95c) and an extent control (Cntle). Large ice anomaly differences exist between the concentration control (Cntlc) and Cntle, complicating the interpretation of the differences when a concentration control is used and thus the Cntlc experiments are not used as a baseline here.

4.0 Conclusions

This study employs an atmospheric global climate model (CCM 3.6) to examine the atmospheric response to observed variations in sea ice during the Summer of 1995, which had the lowest ice extent in the Arctic over the last ~30 years with the exception of 2007. Sea ice was prescribed as ice extent (ocean grid box is either completely

covered or totally ice free) or ice concentration (partial grid box covered in ice allowed) based on monthly varying observations. Fifty-one ensemble members were integrated from April to October 1995 using climatological sea surface temperatures. The control simulation was integrated with global climatological sea ice extent and SSTs. The strongest response was found during the month of August when the ice area is near minimum for the year.

The Arctic displays a local thermal response with increased surface heat fluxes (Sensible + Latent + Longwave) into the atmosphere, warmer SATs and a weak decrease in SLP. The atmospheric response is also characterized by an anomalous high in sea level pressure in the North Pacific and is part of a northward expansion of the summertime subtropical high. The atmospheric response with height is equivalent barotropic and the anomalous high is significant at 200-hPa. There is a significant decrease (increase) of precipitation are along the eastern (northwestern) part of the mean North Pacific storm track. The 500-hPa geopotential height variances and 850-hPa transient eddy heat fluxes indicate enhanced storminess north of the mean storm track and a decrease over the mean storm track in the North Pacific.

Additional climate experiments were conducted to determine the model sensitivity to the location of the sea ice anomalies. When ice reduction is limited to only the Kara Sea, the Laptev-East Siberian Seas, or the Beaufort Sea the atmospheric response patterns for SLP, geopotential height, and precipitation are similar but weaker than when the sea ice is reduced for all the seas, suggesting that the model is sensitive to sea ice anomalies in all three regions. The area of the significant response increases from the Kara to the Beaufort, which is closest to the North Pacific. These results are

analogous to a GCM study by Geisler et al. (1985) where the model Pacific North American (PNA) response pattern (magnitude) is insensitive (sensitive) to the longitude of the tropical Pacific SST anomaly. The August 1995 experiment was repeated using ice concentration anomalies. The atmospheric response is similar though it looks more like a wave train in the Pacific, similar to what Alexander et al. (2004) found for the response during winter to sea ice concentration.

There has been increased interest recently in understanding mechanisms that force and maintain the summertime subtropical highs. In a zonal average, the North Pacific high is strongest in winter when subsidence associated with the Hadley circulation is most vigorous. However, The North Pacific (NP) high is strongest during boreal summer (see Figure 1 of Grotjahn, 2004) and forms to the west of a region with strong thermal contrast between the cool ocean water and the warm North American land mass. Miyasaka and Nakamura (2005) employed a nonlinear spectral primitive equation model at T42 that was driven by zonally asymmetric diabatic heating and demonstrated that the strong surface thermal contrast can explain ~70% of the strength of the subtropical high, consistent with ideas first proposed by Hoskins (1996). Grotjahn (2004) proposed that extratropical storms could provide forcing through transient eddies to maintain the subtropical high. Grotjahn and Osman (2007, Figure 2) present a conceptual picture of how ageostrophic motions arising from developing storms converge at the jet level, leading to sinking motion on the east side of the subtropical high and low-level divergence and southward motion that strengthens the subtropical high. They demonstrate through an observational analysis of daily and monthly SLP that the variability of the NP high is dominated by midlatitude forcing during summer. Some of

the features found in a warm season SLP composite analysis of observations by Grotjahn and Osman (2007) are qualitatively similar to circulation anomalies forced by reduced sea ice in CCM3. They find that SLP is weaker in parts of the Arctic Ocean when the North Pacific high is stronger and a stronger North Pacific high is associated with positive SLP anomalies on the northern flank of the high. Given the similar features outlined above, it warrants further investigation to understand the impact of reduced Arctic sea ice on the variability of the North Pacific high.

The diagnostic analysis suggests that the far field response is not forced directly by the Arctic ice but could rather be a consequence of the local Arctic response, which acts to reduce the flow between the Arctic and the lower latitudes. There may be some parallel with modeling studies of the response to Antarctic sea ice extremes. Hudson and Hewitson (2002) have examined the response to realistic monthly varying sea ice and SST anomalies in the Antarctic. They found that where the sea ice has been reduced and SST exposed, the SAT increases and on the larger scale there is a strengthening and a southward extension of the subtropical high pressure belt. Raphael (2003) has forced the NCAR CCSM with an annual cycle of sea ice maximum and minimum extremes in the southern hemisphere. She finds that reduced ice projects on the positive phase of the southern annular mode (SAM), suggesting either the surface is forcing the SAM (Fyfe 2003) or it is a positive feedback that reduces Southern Hemisphere ice area.

As the Arctic sea ice has continued to decrease in recent years, it is intriguing to consider whether the reduced sea ice may lead to similar large-scale circulation changes in the observations. During August 1995 the observed SLP field displays a

negative anomaly over the Arctic and an anomalous high in the North Pacific (Figure 11a), which is similar to the model response to reduced sea ice. Figure 11b presents a seven year composite of August 500-hPa anomalies based on summers with anomalously low sea ice in the Kara Sea. The anomalous high in the North Pacific bears a strong resemblance to the model response at 500-hPa (Figure 4b). This preliminary analysis showing a similarity between the observations and the model results suggests that modest Arctic sea ice decreases can force circulation changes in the North Pacific.

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FIGURE CAPTIONS

Figure 1. The observed Arctic-wide ice cover ($\times 10^6 \text{ km}^2$) based on ice extent (solid line) and concentration (dashed line) during summer (June-September) over the period 1979-2007 in the HadISST1 $1^\circ \times 1^\circ$ dataset. Ice is defined to extend over a grid square when the ice concentration is 15% or greater. The summer of 1995 had the overall minimum June-September ice extent with the exception of 2007, which was significantly lower.

Figure 2. Evolution of sea ice extent during the boreal summer months of June-August of 1995. Blue (red) squares indicate enhanced (reduced) ice when compared to the monthly mean ice extent (represented by gray+red areas).

Figure 3. (a) Net surface heat flux anomalies (Sensible+Latent+Longwave) felt by the atmosphere and Surface Air Temperature anomalies (b) in response to ice cover of Aug 1995. Dark (Light) pink or blue indicate statistical significance at the 99% (95%) or greater level based on pooled variance t-test. C.I. is 5 W m^{-2} and 0.5°C for heat flux and SAT, respectively.

Figure 4. (a) SLP anomaly response to reduced sea ice during August. Geopotential height anomalies at (b) 500-hPa and (c) 200-hPa. Dark (light) shading indicates statistical significance at the 99% (95%) or greater level based on a pooled variance t-test. The C.I. is 0.5-hPa for SLP and 5 m for geopotential height. This is a polar stereographic view from 40-90N.

Figure 5. Total precipitation anomalies (shaded) are overlaid with contours of mean total precipitation from the control simulation (Cntle) during August. Anomaly magnitudes greater than 0.2 mm day^{-1} are statistically significant at the 95% or greater level based on a pooled variance t-test. C.I. is 0.2 mm day^{-1} for total precipitation anomalies, where red (blue) shading represents positive (negative) anomalies and values between -0.2 and +0.2 are white. C.I. for mean precipitation are 2, 3, 4, 5, 6 and 7 mm day^{-1} .

Figure 6 Storm track variability as indicated by 2-8 day bandpassed a) 500-hPa geopotential height variance, b) 850-hPa $v'T'$, c) 200-hPa $u'v'$ and d) 500-hPa omega variance. The Cntle mean for each quantity is shown by contours and the Sum95e-Cntle anomalies are shown by shaded values. The units are m^2 (a), $\text{m s}^{-1} \text{C}$ (b), $\text{m}^2 \text{s}^{-2}$ (c) and $10^{-4} \text{ Pa}^2 \text{s}^{-2}$ (d). In the Pacific sector statistically significant anomalies have approximate magnitudes greater than 225 m^2 (a), $0.9 \text{ m s}^{-1} \text{C}$ (b), $5 \text{ m}^2 \text{s}^{-2}$ (c) and $6 (10)^4 \text{ Pa}^2 \text{s}^{-2}$ (d).

Figure 7. (a) Mean (Cntle, contours) and anomalous (Sum95e-Cntle, shaded) vertically integrated diabatic heating rate. The path of the transect is shown by a thick black line panel a. (b) Transect through the Arctic into the North Pacific showing total diabatic heating rate anomalies. The units are W m^{-2} in (a) for both shaded and

contoured fields. C.I. is 0.1 K day^{-1} in (b). The total diabatic heating rate is the sum of convective adjustment (DTCOND), solar heating (QRS), longwave heating (QRL), vertical diffusion (DTV) and horizontal diffusion (DTH).

Figure 8. The first column is the geopotential height GCM response to reduced sea ice at 500-hPa (a) and 950-hPa (b). The three right hand side columns show the individual LBM response to transient eddy vorticity fluxes (c-d), transient eddy heat fluxes (e-f) and diabatic heating (g-h). C.I. is 5 m where red (blue) signifies positive (negative) height anomalies. The LBM response has been multiplied by 2.5 to match the magnitude of the GCM response.

Figure 9. Sea ice anomalies for Sum95ke (a), Sum95le (b) and Sum95be (c) where ice is reduced only in the Kara Sea, Laptev-East Siberian Seas and in the Beaufort Sea, respectively. SLP anomalies (d-f) and 500-hPa geopotential height anomalies (g-i) in response to reduced sea ice in individual seas. Dark (Light) shading indicates statistical significance at the 99% (95%) or greater level based on a pooled variance t-test. C.I. is 0.5-hPa for SLP and 5 m for geopotential height.

Figure 10. Sum95c August ice concentration anomalies in % area (a). The Surface Air Temperature anomalies (b) have a C.I. of 0.5 C. The 500-hPa height anomalies (c) for have a C. I. of 5 m.

Figure 11. (a) Observed SLP anomalies in August 1995 in hPa. The mean climatology is based on the years 1968-1996 and the C.I. is 2-hPa. (b) Observed 500-hPa geopotential height composites based on reduced ice in the Kara Sea region. 500-hPa height is in meters. Shading indicates statistical significance at the 95% or greater level based on a pooled variance t-statistic. The years used in this composite are 1979, 1984, 1985, 1994, 1995, 1997, and 2000.